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Tectonic Controls of Mississippi Valley-Type Lead-Zinc Mineralization in Orogenic Forelands

Received: ## ####### #### / Accepted: ## ####### #### / Published online: ## ####### ####

Abstract Most of the world's Mississippi Valley-type (MVT) zinc-lead deposits occur in orogenic forelands. We examine tectonic aspects of foreland evolution as part of a broader study of why some forelands are rich in MVT deposits whereas others are barren. The type of orogenic foreland (collisional versus Andean-type versus inversion-type) is not a first-order control, because each have MVT deposits (e.g. Northern Arkansas, Pine Point, and Cevennes, respectively). In some MVT districts (e.g. Tri-State and Central Tennessee), mineralization took place atop an orogenic forebulge, a low-amplitude (a few hundred m), long-wavelength (100-200 km) swell formed by vertical loading of the foreland plate. In the foreland of the active Banda Arc collision zone, a discontinuous forebulge reveals some of the physiographic and geologic complexities of the forebulge environment, and the importance of sea level in determining whether or not a forebulge will emerge and thus be subject to erosion. In addition to those on extant forebulges, some MVT deposits occur immediately below unconformities that originated at a forebulge, only to be subsequently carried toward the orogen by the plate-tectonic conveyor (e.g. Daniel's Harbour and East Tennessee). Likewise, some deposits are located along syncollisional, flexure-induced normal and strike-slip faults in collisional forelands (e.g. Northern Arkansas, Daniel's Harbour, and Tri-State districts). These findings reveal the importance of lithospheric flexure, and suggest a conceptual tectonic model that accounts for an important subset of MVT deposits—those in the forelands of collisional orogens. MVT deposits occur both in flat lying and in thrust-faulted strata; among the latter group, mineralization postdated thrusting in some instances (e.g. Picos de Europa), but may have predated thrusting in other cases (e.g. East Tennessee).

Keywords Mississippi-Valley-type deposit • foreland basin • collisional orogeny • forebulge • flexural extension

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Introduction

The past quarter century has seen major advances in the understanding of the genesis of Mississippi Valley-type (MVT) lead-zinc deposits (Fig. 1). Until the early 1980s, many workers believed that plate tectonics played no direct role and that MVT mineralization required little more than the presence of platformal carbonates. Recent studies have shown instead that most (though not all) MVT deposits were produced by enormous fluid systems that migrated through foreland basins, driven by gravity from an adjacent orogenic belt (e.g. Garven 1985, Ge and Garven 1992, Appold and Garven 1999, Leach et al. 2001a). Why, then, are some forelands rich in MVT deposits, whereas other forelands are barren? One likely, though admittedly vague explanation, is that foreland basins are not all alike. They form in a variety of convergent tectonic settings at all paleolatitudes, they range from deep, narrow marine flysch troughs to wide, nonmarine clastic wedges, and they evolve through time. Because such factors should have major impacts on regional-scale hydrogeology, they might also be expected to play a role in the genesis, or not, of MVT deposits.

Fig 1 about here.

In this paper, we explore connections between foreland evolution and MVT mineralization. The connection between certain MVT deposits and orogenic forelands has been noted by Leach (1973), Garven (1985), Mitchell (1985), Leach and Rowan (1986), Oliver (1986), Kaiser and Ohmoto (1988), Duane and de Wit (1988), Kesler and van der Pluijm (1990), Bradley (1993), Muchez (2001), Leach et al. (2001a), and many others. Here we scrutinize this relationship.

The search for links between MVT deposits and regional tectonics has been hampered time and again by poor age control. Recent advances in U/Pb and $^{40}Ar/^{39}Ar$ geochronology have revolutionized both orogenic studies and the geologic time scale, the latter being the basis for correlations between orogenic belts and their foreland basins. Mississippi Valley-type deposits are part of the foreland record, but they have been historically difficult to date with isotopic methods because minerals commonly formed in these deposits contain low abundances of useful radioactive isotopes. Despite progress in isotopic and paleomagnetic dating (Leach et al. 2001a), age determinations for MVT deposits still have large uncertainties (e.g. ±10 to 20 m.y). Such error bars span enough time for a “fast” orogeny to run its entire course, from pre-collisional subduction, to syn-collisional mountain building accompanied by hundreds of kilometers of thrust shortening, to post-collisional exhumation and degradation of the foreland basin. Hence it is difficult to relate mineralization to regional tectonic evolution at a level of detail that would be most useful for the problem at hand. Nonetheless, anecdotal information can be pieced together from selected MVT deposits that illustrate specific tectonic controls. We emphasize that not all MVT deposits even formed in foreland tectonic settings; a few seem to be related to extension, and a few formed within thrust belts. Nonetheless, most MVT deposits either are situated in orogenic forelands, or are hosted in rocks that once were located in foreland settings.

In our work, we define MVT lead-zinc deposits as a varied family of epigenetic ores precipitated from dense basinal brines at temperatures ranging between 75° and 200°C, typically located in platform carbonate sequences and lacking genetic affinities to igneous activity (Leach and Sangster 1993). In using this broad definition, we focus on the features that unite a family of ore deposits rather than on the differences that make each MVT district unique. For this reason, we have chosen not to use district names such as "Irish-type", "Alpine-type", or "Viburnum-type".

As used here, an *orogen* is a regional-scale zone of crustal convergence (Fig. 2) that is elevated with respect to its surroundings, and that may be either submarine, or more commonly, subaerial. By this usage, an accretionary prism is an orogen, but a rift shoulder is not. A *foreland* is the region in front of an orogen; in our usage, a thrust belt is not part of the foreland. A *foreland basin* (Fig. 2) is a marine or nonmarine sedimentary accumulation adjacent to an orogen that contains detritus from the orogen, and that subsides in response to orogenic loading. Some workers use *foreland basin* and *foredeep* interchangeably, but we reserve the latter term for an underfilled syncollisional foreland basin—the continental analogue of a trench (Fig. 2A). A *forebulge* (or *flexural bulge* or *peripheral bulge*) (Fig. 3) is a gentle, orogen-parallel swell, typically a few hundred kilometers from the thrust front and a few hundred meters high, that is a distal, elastic response to the same loading that creates a foreland basin. The *far foreland* refers is the region beyond the foreland basin.

Figure 2 near here.

Figure 3 near here.

MVT mineralization in various types of orogenic foreland

Orogenic belts and their associated forelands are produced by interplate or intraplate convergence. Most continental orogens form as the result of arc-continent collision, Andean-type subduction, or basin inversion (Fig. 2). Although all of these tectonic settings involve convergence, they differ in potentially important ways. Factors like the size of the orogenic load, the presence or absence of an attached, subducted slab of oceanic lithosphere, the thermal structure of the flexed plate, and convection patterns in subjacent asthenosphere all have an impact on foreland architecture, and thus might bear on MVT genesis. If foreland-basin architecture does influence the existence or size of the MVT deposits, then tectonic setting could be an important predictive tool for evaluating parts of the world that lack known MVT deposits.

Collisional orogens and their forelands

A *collisional orogeny* is one in which subduction of oceanic lithosphere leads to impact between two non-subductible objects, such as continents, microcontinents, arcs, and oceanic plateaux. Collisions follow from a number of plate geometries, but the most common, and the one that pertains to MVT deposits, is collision between passive margin and an arc. The schematic model in Figure 3 is based on Neogene examples in Timor, New Guinea, and Taiwan, and older examples including two associated with MVT deposits, the Taconics and Ouachitas of North America.

Continental breakup by rifting leads to thermal subsidence of a matched pair of passive margins adjacent to a widening ocean. Eventually, plate reorganization leads to subduction somewhere in this ocean basin, and an arc and one of the margins begin to converge. Few if any passive margins appear to be immune to this fate: a survey of Phanerozoic examples disclosed that all collided with an arc within a few hundred million years of forming (Burke et al. 1984). Seafloor feeding into a subduction zone passes over a gentle forebulge, goes down the outer trench slope where it is cut by normal faults, is buried by trench sediments, and finally is carried beneath the frontal thrust, either to be offscraped, underplated, or subducted into the mantle (Fig. 3A). With continued plate convergence, a point on the passive margin platform eventually reaches and then migrates through the axis of the forebulge, and then continues across a zone of active normal faulting on the outer slope of the foredeep, and into the foredeep axis. In collisions that take place at low latitudes (i.e. where a carbonate platform is involved), the resulting upward-deepening succession is striking and unmistakable (e.g. Bradley & Kusky 1986; Robertson 1987; Sinclair 1997). Typically, the forebulge unconformity is overlain by shallow-marine carbonates that give way upward to carbonate turbidites, shales, and finally, orogenically derived siliciclastic turbidites (flysch). Depending on the geometry of thrusting, the sedimentary cover of the lower plate is either subducted with its basement, or is detached and becomes part of the growing thrust belt. Plate convergence perpetuates this scheme in a more-or-less steady-state fashion (Hoffman 1987). Eventually, when convergence slows and then stops, the various paleogeographic elements in the foreland are buried by an upward-shallowing siliciclastic sequence (molasse).

Many accounts of foreland-basin evolution end here, but to understand MVT genesis, the aftermath of collision cannot be ignored. Erosion of a recently formed mountain belt removes part of the load that created the accommodation space for the foreland basin (Beaumont et al. 1993). Erosional unloading results in uplift of both the mountains and the proximal foreland basin (Fig. 3C). Extensional collapse of an orogenic hinterland (e.g. Dewey 1988; de Boorder et al. 1998) has a similar unloading effect (Fig. 3D), with erosion acting in concert with normal faulting to reduce the orogenic load. Either process *increases* topographic relief and thus may help to drive MVT fluids toward the adjacent foreland (Garven 1985; Bethke and Marshak 1990; Garven et al. 1993; and Appold and Garven, 1999). With continued erosion, a foreland basin eventually becomes so degraded as to lose its hydrological integrity, and thus lose its potential to transmit the enormous volumes of fluids required of MVT systems.

Some MVT districts are clearly related to collisional orogenic forelands. The most straightforward example is provided by Ozark MVT deposits that include the Southeast Missouri (Viburnum Trend and Old Lead Belt), Northern Arkansas, Tri-State and Central Missouri districts, which lie in the Ouachita foreland (Fig. 4). The Ouachita orogen is a fold-thrust belt that formed during the Mississippian to Pennsylvanian collision between the passive margin of Laurentia and an accretionary wedge in front of a north-facing arc. Collision followed subduction of an ocean of unknown width (e.g. Viehle 1979; Lillie et al. 1983; Houseknecht 1986). Breakup of Rodinia during the Late Precambrian led to the formation of the Ouachita passive margin by Cambrian time (Thomas 1991). The passive margin endured about 200 m.y. In thrust sheets in the Ouachita Mountains, Cambrian to Lower Mississippian slope and rise facies are overlain by Lower Mississippian (late Meramecian, ~336-331 Ma)[1] flysch. The flysch is interpreted to record arrival of the distal continental margin at a trench. The frontal part of the orogen and adjacent flat-lying rocks to the north define the Arkoma foreland basin, filled by as much as 8 km of flysch and molasse. These strata were deposited during the Atokan (Early Pennsylvanian, ~314-311 Ma) and Desmoinesian (mid-Pennsylvanian, ~311-305 Ma) (Houseknecht 1986; Sutherland 1988). The Arkoma Basin laps to the north onto the Ozark dome, a tract of mainly Cambrian to Pennsylvanian carbonate-dominated platform facies that host the MVT deposits. The older carbonates were deposited first along the Laurentian passive margin, facing an open ocean, whereas the youngest carbonates were deposited along the cratonic margin of the advancing foredeep. The MVT deposits in the Ozark region are located on the area of the forebulge, and some orebodies are located along flexure-induced normal faults. The deposits formed after collision. These points are elaborated in subsequent sections. The single most important point illustrated by the Ozark MVT deposits is that they are related to a simple arc-passive margin collision; no subsequent orogenies complicate matters.

Figure 4 near here.

Andean-type orogens and their forelands

Andean orogens form above continental-margin subduction zones in which the upper plate is in compression, as manifested by crustal thickening along the arc and by development of a thrust belt and foreland basin in the backarc (Fig. 2B) (Dewey 1980). Although some early plate-tectonic models suggested otherwise, Andean-type arcs appear not to form directly by initiation of subduction along passive margin, but instead by one or more collisions, the first involving a passive margin and an arc, followed by subduction flip (Burke et al. 1984).

MVT deposits of the Canadian Rockies and its foreland show a clear relationship to Andean-type orogeny. Late Proterozoic rifting followed by thermal subsidence during the early Paleozoic (Bond and Kominz 1984) produced a westward-thickening carbonate platform flanked to the west by deep-water facies. Cambrian to Late Devonian passive-margin carbonates host the MVT deposits. Near the end of the Devonian, an influx of flysch from outboard sources most likely records collision of a Devonian arc (Smith et al. 1993). This orogenic clastic sequence never prograded very far to the east; on the miogeocline, mixed carbonate and clastic platform deposition lasted into mid-Jurassic. From then until the Paleocene, an Andean-type thrust belt (Canadian Rockies) and its

[1] Carboniferous time scale based on Europe-North America correlations summarized in Harland et al. (1990), calibrated to the numerical time scale of Menning et al. (2000).

foreland basin (Western Interior basin) advanced cratonward (McMechan and Thompson 1993). The MVT deposits of western Canada occur in both the undeformed foreland (Pine Point) and in the thrust belt (Robb Lake and Monarch-Kicking Horse). The flat-lying rocks of the Pine Point district are located near the cratonward edge of the Western Interior foreland basin. The ore deposits are hosted in paleokarst in a Middle Devonian carbonate barrier complex. Rb/Sr dating of sphalerite has yielded isochron ages of 361 ± 13 Ma and 374 ± 21 (Nakai et al. 1993; J. Brannon personal communication in Symons et al. 1998a). These results are not much younger than the host rock, and may date Devonian clays entrapped in much younger sphalerite (Symons et al. 1998). Paleomagnetic dating at Pine Point indicates a more plausible latest Cretaceous to Paleocene age (71 ± 13 Ma) (Symons et al. 1993), implying mineralization in a distal Andean-type foreland. As will be discussed later, the MVT deposits in the thrust belt also have yielded Late Cretaceous and Eocene paleomagnetic ages (Symons et al. 1998a; Smethurst et al. 1999), in support of the younger age for Pine Point.

Inversion-type orogens and their forelands

Orogens formed by inversion of sedimentary basins (Fig. 2C) comprise a third category known to be associated with MVT deposits. These orogens do not form upon closure of an ocean basin by subduction, but within continental crust. To call such orogens “collisions” is a misnomer, because the two converging objects are already juxtaposed from the start. The example most pertinent to MVT deposits is the Pyreneean Orogen of southern Europe. During the opening of the North Atlantic, as the Americas moved away from Europe and Africa, Iberia moved as an independent block. From Permian to Late Cretaceous, a series of rift basins subsided between Europe and Iberia (e.g. Puigdefabregas and Souquet 1986). From late Santonian to Miocene, oblique, and then orthogonal convergent motions inverted these basins, and formed the Pyreneean Orogen (Puigdefabregas et al. 1992). The Pyrenees are a doubly vergent thrust belt flanked north and south by foreland basins and by synorogenic MVT deposits (Fig. 5). For the Cevennes MVT district in the northern foreland, paleomagnetic dating places the time of mineralization between about 60 and 50 Ma (late Paleocene to early Eocene) (Lewchuk et al. 1998), a time of major foreland-basin infilling. The paleomagnetic age agrees with recent isotopic ages obtained on fluorite associated with some of the lead-zinc ores in Cevennes (Leach at al. 2001b). Similarly, for the much smaller Maestrat MVT district in the southern foreland, U/Pb dating of ore-stage calcite suggests that mineralization took place at about 63 Ma (early Paleocene) (Grandia et al. 2000). These age determinations show that mineralization in both forelands took place during Pyreneean mountain building.

Fig. 5 (new) about here.

Conclusions on tectonic setting

We are led to conclude that, across the spectrum of orogens, tectonic setting is *not* a first-order control on MVT-forming processes. This is not to say that tectonic setting is unimportant (for example, it might bear on the size of deposits, or on local structural controls), only that it does not hold much promise as a first-order exploration guideline in frontier areas.

MVT mineralization and forebulges

Forebulges occur next to various kinds of vertical loads on the lithosphere: continental ice sheets, hotspot volcanoes, and, as already mentioned, orogenic belts (Quinlan and Beaumont 1984). Orogenic forebulges form whether the load is advancing or stationary; as will be shown, both cases are pertinent to MVT deposits.

As a passive margin approaches an arc with which it is destined to collide, the attached ocean floor is first subducted. Ideally, bending of the subducting lithosphere produces a forebulge (Fig. 3), which migrates like a gentle wave across the downgoing plate, in advance of the plate boundary itself. The bulge is not directly related to compression, but rather to the flexing of an elastic plate over a viscous substrate, due to vertical loading. During the earliest phases of a collision, the passage of the continental slope and rise through a forebulge is unlikely to leave an obvious record, simply because a given location remains in deep water before, during, and after the forebulge. But as convergence continues, the miogeoclinal platform reaches the bulge, where even a few meters of vertical motion can have major effects on the sedimentary record. As it approaches the forebulge, a given location on the platform shoals and may emerge above sea level. (Note: Eustatic sea level is no doubt an important factor in determining whether or not a forebulge *unconformity* develops.) With continued plate convergence, this same location then begins to subside as it passes the crest of the bulge and approaches the foredeep axis. The clearest modern example is at Kepulauan Aru, the modern emergent forebulge related to collision between the Banda Arc and the passive margin of northern Australia (Bradley and Kidd 1991) (Figs. 6 and 7). The axis of this ~75-km-wide island chain lies about 100 km from the deformation front and consists of a deeply karsted Neogene limestone terrane with relief as great as 240 m, surrounded by shallow carbonate seas. Significantly, a forebulge is *not* developed in the expected position to the southwest of Kepulauan Aru (Fig. 6). Thus, it is not a foregone conclusion that a forebulge (or forebulge unconformity) will always form in a collisional foreland.

Figure 6 near here.

Figure 7 near here.

Forebulge unconformities are key features in some MVT systems. Certain MVT deposits in the Ozarks, the Appalachians, and Spain occur along unconformities that formed at forebulges—or more cautiously stated, at the right place and at the right time to have been caused by forebulges. Such unconformities occur immediately below the top of shallow-marine carbonate platform sequences, and are capped by an upward-deepening sequence of shallow marine carbonates, deep marine carbonates, black shale, and siliciclastic turbidites that together constitute the foredeep succession. Forebulge unconformities (Jacobi 1981) are an important ore control at three places in the Appalachians: Daniel's Harbour (Fig. 8), Friedensville, and the East Tennessee district (Bradley 1993). Significantly, the age of the unconformity varies by a few million years along strike, so eustatic sea-level fall cannot be the sole explanation for emergence of the shelf. The unconformity is impressive. In Virginia, for example, erosional relief is as great as 140 m, and sinkholes and caves extend to 65 m below the unconformity (Mussman and Read 1986). In Newfoundland, similarly, Knight et al. (1991) reported erosional relief up to 50 m and karst features to depths of 120 m.

In the Ozarks (Fig. 9), the youngest carbonate rocks (Chesterian to Morrowan) record four episodes of erosion that are broadly syncollisional (Brockie et al. 1968; Sutherland 1988). Timing and location together suggest that at least some of these unconformities were produced at a forebulge in front of the advancing orogen (Kaiser and Ohmoto 1988).

Finally, in the Picos de Europa district of Spain, MVT mineralization (e.g. Aliva deposit, Fig. 5) is hosted in Namurian to Westphalian platformal carbonates (Gomez-Fernandez et al. 2000). The carbonates are depositionally overlain by Stephanian flysch deposited in a Hercynian foreland basin. The carbonates and flysch are separated by an unconformity that we attribute to a Hercynian forebulge.

Figure 8 near here.

Figure 9 near here.

The distinction between a forebulge and a forebulge unconformity is straightforward, but merits emphasis because it pertains to the position of certain MVT deposits. A forebulge is a physiographic feature, whereas a forebulge unconformity is a stratigraphic surface that may or may not still be located on the physiographic bulge. The Central Tennessee MVT district (Gaylord and Briskey 1983) is situated on the Nashville Dome, one sector of the Appalachian forebulge (Quinlan and Beaumont 1984). In contrast, the Ordovician-hosted MVT deposits in the Appalachian thrust belt (e.g. East Tennessee) lie along a diachronous unconformity that formed as the passive margin passed over this forebulge during the Ordovician. Rocks that had been on the forebulge eventually ended up in the thrust belt.

A final issue that has not yet been addressed in MVT studies has to do with the migration, or not, of forebulges *after* orogenesis. That forebulges migrate in advance of *moving* loads is self evident from their existence on the seaward flank of most deep-sea trenches, where plate convergence is measured in centimeters per year. When a tectonic load stops moving, the expected response depends on the rheology of the lithosphere. If the lithosphere is perfectly elastic, a stationary load of constant size is flanked by a stationary forebulge. On the other hand, if the lithosphere is viscoelastic (i.e., it relaxes stress), the forebulge migrates toward the load (Beaumont et al. 1993). When a stationary orogenic load shrinks, as it inevitably does due to erosion, the forebulge should also migrate toward it (Beaumont et al. 1993). Forebulge migration toward a stationary orogen is one explanation for an apparent southward younging of MVT mineralization from Central Missouri to Northern Arkansas in the Permian (303±17 Ma in the north, 265±20 in the south) (Leach et al. 2001a). Central Missouri now lies north of the axis of the Ozark forebulge in the Ouachita foreland (Fig. 9), but could have been on the forebulge axis at the time of mineralization. This interpretation provides a plausible mechanism for the northward flow of mineralizing fluids from the Arkoma Basin past the present forebulge axis.

Structural control of MVT mineralization in orogenic forelands

Tectonic structures control the location of MVT mineralization in many orogenic forelands. Such structures fall into two main categories: (1) synorogenic, and (2) other. Synorogenic foreland faults are important in some MVT systems. Bending of subducting lithosphere through angles greater than about 1.5° generally results in normal faulting of the convex outer surface of the flexed plate (Jacobson et al. 1979). This process, termed “flexural extension” by Bradley and Kidd (1991), takes place in both oceanic and continental lithosphere, in spite of an overall setting of plate convergence. Flexurally induced normal faults localized MVT mineralization in the Appalachian, Ouachita, and Carpathian forelands. Location, orientation, and timing link normal faulting to collisional orogeny. In the Ouachita foreland (Fig. 4), a prominent set of east-west-striking, orogen-parallel normal faults has long been recognized as syncollisional; growth strata in the foreland basin show that faulting was Atokan in age (Houseknecht 1986). Normal faults of this set localized mineralization in the Northern Arkansas MVT district (e.g. St. Joe, Tomahawk Creek, and Rush Creek areas; McKnight 1935). Similarly, in the Taconic foreland of Newfoundland, MVT mineralization at Daniel's Harbour is localized along orogen-parallel normal faults that influenced Ordovician platformal facies near the onset of collision (Lane 1984; Knight and James 1987; Knight et al. 1991). The Cracow-Silesian MVT deposits of Poland likewise are associated with Tertiary normal faults in the Carpathian foreland basin (Symons et al. 1995). Mitchell (1985) related the normal faults that control mineralization in the Irish MVT district to flexural extension in the Variscan foreland. As summarized by Hitzman (1999), however, normal faulting took place during the Visean (mid-Mississippian), whereas Variscan shortening is Westphalian and younger.

The properties of flexurally induced normal faults (Bradley and Kidd 1991) are relevant to MVT exploration. Normal faults formed by flexural extension typically strike parallel to the orogenic front, are downdropped toward the orogen, and cause only minor landward stratal rotations. Whereas stratigraphic throws up to a few hundred meters are typical, the biggest known faults have throws in excess of 2000 m. The major normal faults in the Taconic foreland are typical: they are spaced 10 to 20 km apart and pervade a ~100-km-wide swath in the foreland. Recently, Hudson (2000) has shown that flexural extension in the Ouachita foreland resulted in a complex orthorhombic network of faults that included not only normal faults, but also strike-slip and reverse faults. Being much younger than rift-related faults, fault related to flexural extension cut the entire carbonate section and typically die out upward within the foreland-basin fill. Hence, these faults can be expected to offset paleoaquifers, setting up conditions for fluid mixing, one of the keys to precipitation of MVT ores.

In addition to the normal faults described above, the Ouachita foreland is also cut by a number of synorogenic strike-slip faults that are oblique to orogenic strike (Fig. 4) (Hudson 2000). Faulting appears to have begun as early as Late Mississippian (Chesterian) time, when collision was beginning and the deformation front still lay some 200 km to the south. In the Tri-State lead-zinc district (Fig. 4), movement along the northeast-striking Miami fault system during Chesterian deposition is evident from the presence of an anomalous thickness of strata of that age within a downdropped block along the fault zone (Brockie et al. 1968). Analogous faults cut the modern collisional forebulge at Kepulauan Aru (Fig. 7).

Some MVT-controlling structures are only coincidentally related to the foreland in which they occur. MVT mineralization at Polaris and nearby deposits in the Canadian Arctic is controlled by a large-scale structure that has no genetic connections with the Ellesmerian collisional foreland in which the deposits formed (Fig. 10). Ellesmerian orogenic history took place in three phases: Late Precambrian rifting, Cambrian-Ordovician passive-margin subsidence, and Silurian-Devonian southward advance of a collisional foreland basin (e.g. Trettin et al. 1991). Polaris is hosted in Ordovician platform carbonates of the passive margin. A Late Devonian paleomagnetic age (Symons and Sangster 1992) places MVT mineralization near the end of Ellesmerian orogenesis. Although the carbonate platform can be traced for 2000 km from northeastern Greenland to the western Canadian Arctic, MVT mineralization is concentrated in a small area where the platform is interrupted by the Boothia Uplift, a 1000-km-long, north-south belt of Late Silurian-Early Devonian deformation that strikes at a high angle to the continental margin (Kerr 1977a, 1977b). Okulitch et al. (1986) suggested that the Boothia Uplift was a far-field effect of Caledonian collision between Baltica and Laurentia, somewhat akin to the present-day Tien Shan ranges in the far foreland of the India-Asia collision (Molnar and Tapponier 1975). Whatever its cause, the pre-existing Boothia Uplift likely played a role in focusing mineralizing fluids during Ellesmerian orogenesis.

Figure 10 (new) about here.

In Southeast Missouri (Fig. 4), some 400 km cratonward of the Ouachita thrust front, high-angle faults are one of several controls of MVT mineralization, which has been dated paleomagnetically as Early Permian (Symons et al. 1998b). Clendenin et al. (1993) demonstrated that faulting was a manifestation of Cambrian rifting along the southern margin of Laurentia. Some faults were reactivated in late Paleozoic in the Ouachita foreland (Kaiser and Ohmoto 1988; Clendenin et al. 1989), close to the time of MVT mineralization. Thus, events formed on either end of a single Wilson Cycle combined to form ore-controlling structures in Southeast Missouri.

Importance of plate convergence

Plate convergence is what drives an orogen forward, causing its harbingers, the forebulge and domain of flexural extension, to migrate. Thus, plate convergence thus feeds a carbonate platform through a karst factory, in the manner of a conveyor belt (Bradley 1993). More convergence creates more normal faults and more karst. A palinspastic restoration of the Taconic foreland shows how this worked in the Appalachians (Fig. 11). At the close of Taconic collision, it would have been possible to trace the forebulge unconformity some 300 km across strike, even though the topographic forebulge was perhaps only 50 to 100 km wide. Theoretically, the rate of plate convergence should also control how long it takes for a particular location on the passive margin platform to migrate through the forebulge. Other factors being equal, slower convergence rates should correspond to more time on the forebulge, and thus, deeper karst erosion.

Figure 11 near here.

Overfilled versus underfilled foreland-basins during MVT mineralization

At any given time in its evolution, a foreland basin can be categorized as underfilled versus overfilled. Typically, foreland basins are underfilled at the beginning of orogenesis, and eventually fill with sediment as plate convergence grinds to a halt and the balance shifts between rates of tectonically driven subsidence and sedimentation (Fig. 12). The modern Timor Trough is an example of an underfilled basin and the New Guinea foreland basin is a example of an overfilled one (Fig. 6). When plate convergence stops, both the mountain belt and its foreland basin rise as a consequence of erosion (Fig. 3C) or, alternatively, due to extensional orogenic collapse (Fig. 3D). Regardless of the mechanism, orogenic unloading seems to be an ideal way to create an elevated recharge area that would connect with essentially undeformed sedimentary aquifers in the depths of the foreland basin, and thence to discharge at places hundreds of kilometers away on the craton. Indeed, this is the configuration that Garven (1985) successfully modeled in his Pine Point study. The Ozark MVT deposits provide the best example of mineralization related to an overfilled foreland basin. Mineralization, dated as Permian by several methods in several districts (see review by Leach et al. 2001a), took place long after the transition from marine to nonmarine conditions and after all significant contractional deformation. Although Permian strata are not preserved in the Arkoma Basin, modern analogues such as New Guinea (Fig. 6) imply that the topographic surface sloped from the orogen toward the foreland (Fig. 9).

Figure 12 near here.

The question must remain open as to whether or not MVT mineralization can take place during the underfilled stage of foreland-basin development.

MVT deposits in thrust belts

Some MVT deposits are in flat-lying strata, whereas others are in deformed host rocks within orogenic belts. For each deposit of the latter group, the question is whether it formed in strata that were flat-lying at the time, or in already-deformed host rocks (Fig. 13). Both cases exist.

Figure 13 near here.

MVT mineralization in East Tennessee has long been regarded as predating thrust-related deformation, because, at Flat Gap and Mascot-Jefferson City, sphalerite sands dip parallel to the dip of the now-inclined bedding in host dolomites (Kendall 1960; Hoagland et al 1965; Hill et al. 1971). If the sphalerite grains are indeed clastic, mineralization predates deformation. Alternatively, however, the sphalerite sands could have formed by grain-by-grain replacement of pre-tectonic carbonate sands (Symons and Stratakos 2000); such replacement could have happened before, during, or after deformation. Thus, the timing of mineralization with respect to thrusting in East Tennessee is debatable.

On the other hand, geologic evidence shows that the Picos de Europa MVT district in the Hercynian Orogen of northern Spain (Fig. 5) formed where it now is, in a thrust belt. As noted in the discussion of forebulges, these deposits are hosted by Carboniferous carbonates. The host strata were imbricated by Hercynian thrusts of latest Carboniferous (Stephanian) age, and these thrusts are cut, in turn, by east-west high-angle faults of Permian age (Gómez-Fernandez et al. 2000). MVT mineralization is concentrated along these younger faults and thus postdated thrusting.

A similar mineralization interpretation is emerging for MVT deposits in the Canadian Rockies thrust belt. The Monarch and Kicking Horse deposits are hosted in Cambrian carbonates; the Robb Lake deposit is in Upper Silurian to Middle Devonian carbonates. A Late Cretaceous (but pre-84 Ma) paleomagnetic age for the Kicking Horse deposit shows that mineralization was broadly coeval with Laramide thrust-related deformation. An Eocene paleomagnetic age for the Robb Lake deposit (Smethurst et al. 1999) shows that mineralization postdated the main thrust-related folding. We conclude that these deposits formed in imbricated carbonate thrust panels within a mountain belt—a very different hydrologic setting than that pictured for MVT deposits in forelands (e.g. Garven 1985).

Two Appalachian MVT deposits, Daniel’s Harbour and Friedensville, fall between these two end-member cases. These deposits appear to have formed beneath thrust sheets near the orogenic front, but in essentially undeformed Ordovician strata. Both locations were buried by thrusts during the Taconic Orogeny and were last situated in a foreland basin immediately before arrival of the thrust sheets, in Ordovician time (Bradley 1993) (Fig. 11). Paleomagnetic dating of the Daniel’s Harbour deposit suggests a Middle Devonian mineralization age (Pan and Symons 1993), corresponding to the second Paleozoic orogeny in the Appalachian foreland, the Acadian (Cawood and Williams 1988). Apparently, some 60 million years elapsed between initial tectonic burial and eventual mineralization..

Summary and closing comments

Why, then, do MVT deposits occur in some orogenic forelands, but not others? Not all controls are tectonic, but even if they were, there still would be no simple answer. For a deposit to form, a number of different factors must come together: e.g. suitable host rocks, ground preparation, basinal brines, and mechanisms to drive these brines and focus them at the ore site (Leach and Sangster 1993). However, for exploration or mineral assessment in frontier areas, the present study does suggest some useful guidelines.

- Because most MVT deposits are in carbonate rocks and most thick carbonate successions form at low latitudes, a pre- to syn-orogenic drift history through low latitudes is a prerequisite (Leach et al. 2001a).
- The type of orogenic foreland (collisional, Andean-type, inversion-type) is not critical to the presence or absence of MVT deposits. But these foreland systems do differ in certain respects such as basin geometry and structure, which could influence exploration strategies.
- Forebulge unconformities formed during collisional orogeny are a prime target; they are recognized by their place in the stratigraphic succession (e.g. Fig. 8). By this criterion, the Cenomanian-Turonian unconformity in the collisional foreland of Oman (Robertson 1987) should be a good target for MVT exploration.
- Syncollisional normal and strike-slip faults in collisional forelands host MVT mineralization in some districts and thus should be exploration targets in foreland systems.
- The timing of MVT mineralization with respect to a collisional orogeny (syn- or post-) is directly relevant to MVT genesis and also could bear on exploration. The Ozark MVT deposits reveal that mineralization can take place on a massive scale, a few tens of millions of years after orogenesis. At the time of mineralization, the foreland basin was overfilled and the topographic gradient sloped away from the orogen.
- MVT deposits occur both in flat-lying and in thrust-faulted strata; both types of terrane are prospective. For the latter group, the question is whether mineralization predates or postdates thrusting.

MVT deposits related to arc-passive margin collision (e.g. those of the Ozark region) warrant special mention because this type of orogen provides all of the key ingredients for mineralization, in a single chain of events. Thus we offer a model for MVT genesis *in this settin*g (Fig. 12), in which we describe events from the point of view of an observer on the arc being approached by a passive margin (the choice of reference frame is arbitrary). Suitable host carbonates are deposited for many tens of millions of years in a seaward thickening miogeoclinal wedge along a thermally subsiding passive margin. Subduction begins somewhere in the ocean basin, and the arc and passive margin begin to converge. During the earliest stages of collision, migration of the platform through the forebulge causes gentle uplift, emergence, and karstification. Over time, with continued plate convergence, many tens to hundreds of kilometers of platform pass through the forebulge, preparing ground for mineralization over a much broader across-strike area than might otherwise be possible. Flexure of the passive margin beneath the orogen drives subsidence of the foredeep, which fills with siliciclastic strata. Flexure also causes extension of the carbonate platform, forming normal faults that may later serve to focus basinal fluids. As collision slows and then stops, what had been an underfilled foreland basin can finally fill with sediment (Fig. 12B). Erosion removes part of the orogenic load, and both the lower plate (with its foreland-basin cover strata) and the surviving mountains consequently rebound (Beaumont et al. 1993). Recharged by rain at or near the mountain front, a gravity-driven regional fluid system forms, and large volumes of fluids are thereby delivered to distant sites of mineralization (Fig. 12B). Eventually, some factor or combination of factors (erosion, climate change, deformation) interrupts the regional fluid system and mineralization stops.

Given the existence of MVT deposits in settings other than arc-passive margin collisions, it is clear that the ingredients for mineralization can be created and put together in more than one way. In this light, among the most intriguing MVT systems are those like the Canning Basin district (Christensen et al. 1995) and Nanisivik (Symons et al. 2000), that appear to have formed in extensional settings. Equally interesting are those MVT deposits, like Robb Lake and the Picos de Europa district, that appear to have formed within thrust belts. Other tectonic models are needed for MVT genesis in these settings.

Acknowledgments

We thank Grant Garven, Gerry Stanley, Alex Brown, and Rich Goldfarb for reviewing the manuscript. Collaborations with Don Sangster, Mike Lewchuk, David Symons, Henri Rouvier, and Jean-Claude Macquar have greatly influenced us over the years. Rod Randell, Tom Lane, and Antonio Alonso kindly led us through the Polaris, Daniel's Harbour, and Reocin deposits, respectively.

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FIGURE CAPTIONS

Figure 1. World map of MVT deposits. Those discussed in this paper are shown with black circles. North America: (1) Washington Land, (2) Polaris, (3) Eclipse, (4) Nanisivik, (5) Reef Ridge, (6) Gayna, (7) Bear-Twit, (8) Godlin, (9) Pine Point, (10) Esker, (11) Robb Lake, (12) Monarch-Kicking Horse, (13) Giant, (14) Metaline, (15) Upper Mississippi Valley, (16) Central Missouri, (17) Southeast Missouri (Old Lead Belt, Viburnum Trend, Indian Creek), (18) Tri-State, (19) Northern Arkansas, (20) Central Tennessee, (21) East Tennessee (Mascot-Jefferson City, Flat Gap), (22) Austinville, (23) Friedensville, (24) Gays River, (25) Daniel's Harbour. South America: (26) San Vincente, (27) Vazante. Eurasia: (28) Ireland (e.g. Navan, Lisheen, Galmoy), (29) Picos de Europa, (30) Reocin, (31) Maestrat, (32) Cévennes, (33) Sardinia, (34) Alpine district, (35) Cracow-Silesia, (36) Irankuh district. Africa: (37) El-Abad-Mekta district, (38) Bou Grine, (39) Pering-Bushy Park. Australia: (40) Sorby Hills, (41) Coxco, (42) Lennard Shelf (e.g. Cadjebut, Blendvale, Twelve Mile Bore).

Figure 2. Collisional, Andean, and transpressional orogens compared. (A) Arc-passive-margin collision, based on Neogene examples from Timor, New Guinea, and Taiwan, and various older examples including the Taconic and Ouachita orogenies. (B) Andean-type orogen, based on the modern Andes and the Late Cretaceous-Paleocene Laramide system of western North America. Convecting asthenosphere contributes to foreland subsidence on a broad, regional scale (Mitrovica et al. 1989), setting this type of foreland system apart from others. (C) An inversion-type orogen flanked on both sides by thrust-loaded foreland basins, based on the Pyrenees (Muñoz 1992).

Figure 3. Sequential model for arc-passive margin collision. A leads to B which leads to either C or D. C and D are alternative explanations for late to immediately post-orogenic uplift.

Figure 4. Map of the Ouachita Orogen, Arkoma foreland basin, and the corresponding forebulge, known as the Ozark Dome. Faults in the foreland are mainly synorogenic and include (1) normal faults that parallel the thrust front, and (2) strike-slip faults at an oblique angle to the thrust front. Faults in Southeast Missouri originated as Cambrian rift structures and were reactivated during Ouachita collision (Clendenin et al. 1993). Note the patchy distribution of MVT deposits in the foreland. Central Missouri MVT district lies just off the map to the north. Modified from Bradley and Kidd (1991).

Figure 5. Generalized geologic map of the Iberian Peninsula and southwestern France, showing the occurrence of MVT deposits in both northern and southern forelands of the Pyreneean Orogen.

Figure 6. Generalized geologic map of collision zones involving the northern passive margin of Australia. Ongoing collision with the forearc of the Banda Arc has produced a collisional foreland basin, Timor Trough, which can be traced to the west into an oceanic subduction zone, the Java Trench. Timor Trough is an underfilled foreland basin. Kepulauan Aru (Fig. 6) is interpreted as an

emergent forebulge. Sahul Rise appears to be a drowned version of Kepulauan Aru that was exposed in the Pleistocene (Veevers and van Andel 1967). Note the absence of a bathymetric forebulge in the ~600-km gap between the two sites. New Guinea represents an older arc-passive margin collision that has nearly ground to a halt. What had been an underfilled foreland basin has now filled with fluvial sediments and has a surface that grades to the south, away from the mountain front. Thus, present-day southern New Guinea resembles the Ouachita foreland at the time of MVT mineralization.

Figure 7. Geologic map of Kepulauan Aru (Aru Islands), Indonesia, adapted by Untung (1985) from Indonesian literature. This is the clearest modern example of an emergent forebulge on a carbonate platform in a collisional foreland. The deformation front lies about 100 km to the west. Platformal strata as old as middle Miocene are being eroded. The highest elevation is 240 m. NW-trending drowned river valleys are interpreted as antecedent drainages that existed prior to forebulge uplift. The NE-striking faults, cutting rocks as young as Pliocene, are analogous to structures like the Miami fault, which controlled mineralization in the Tri-State MVT district.

Figure 8. Generalized stratigraphic section of western Newfoundland, in the vicinity of the Daniel's Harbour (Newfoundland Zinc) MVT deposit, adapted from James and Stevens (1982). The succession below the forebulge unconformity was deposited on a passive-margin platform at relatively low rates of subsidence. The unconformity has erosional relief of about 50 m. Karst features host MVT mineralization. Above the unconformity, an upward-deepening carbonate sequence gives way to siliciclastic turbidites of the Taconic foreland basin. Note the order-of-magnitude difference between subsidence rates below and above the unconformity.

Figure 9. Schematic north-south cross-section through the Ouachita Orogen and its foreland, representing conditions in Permian time, after plate convergence had ended. The line of section extends slightly beyond the north-south limits of Figure 4. MVT deposits are projected onto the section to indicate their across-strike position relative to the orogen, foreland basin, and forebulge.

Figure 10. Generalized geologic map of the Canadian Arctic showing the position of the Polaris MVT deposit at the juncture of the Boothia Uplift and Ellesmerian foreland basin. Boothia Uplift is an intracontinental thrust belt that may have been activated the distant collision between Greenland (then part of Laurentia) and Baltica (Okulitch et al. 1986).

Figure 11. Schematic cross-section through the Taconic foredeep at the end of Ordovician orogeny. The distribution and widths of paleogeographic elements are based on the foredeep in New York (Bradley 1989), but are representative of all transects through the foredeep along the length of the Appalachians.Positions of Ordovician-hosted MVT deposits are projected onto the section relative to paleogeographic elements when plate convergence ceased. Two locations where MVT deposits eventually formed (Friedensville and Daniel's Harbour) were tectonically buried beneath Taconic thrust sheets during Ordovician time.

Figure 12. Block diagrams showing foreland evolution. (A) During plate convergence, submarine thrust belt loads the passive margin, thereby forming the foreland basin, extensional domain, and forebulge. Plate convergence continually causes these features to migrate across the foreland plate. Foreland basin remains underfilled because the depocenter migrates. Barite mineralization along foredeep axis is based on examples from the Ouachitas (Maynard and Okita 1991). (B) Plate convergence has ceased and foreland basin has filled with sediment, creating hydrologic conditions favorable to MVT mineralization. This is the situation corresponding to mineralization in the Ozark region.

Figure 13. Two alternative sequences for the genesis of MVT deposits in thrust belts. In (A), deposits form in flat-lying strata and are subsequently deformed. In (B), deformation of host strata precedes mineralization.

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Fig. 1, Bradley & Leach

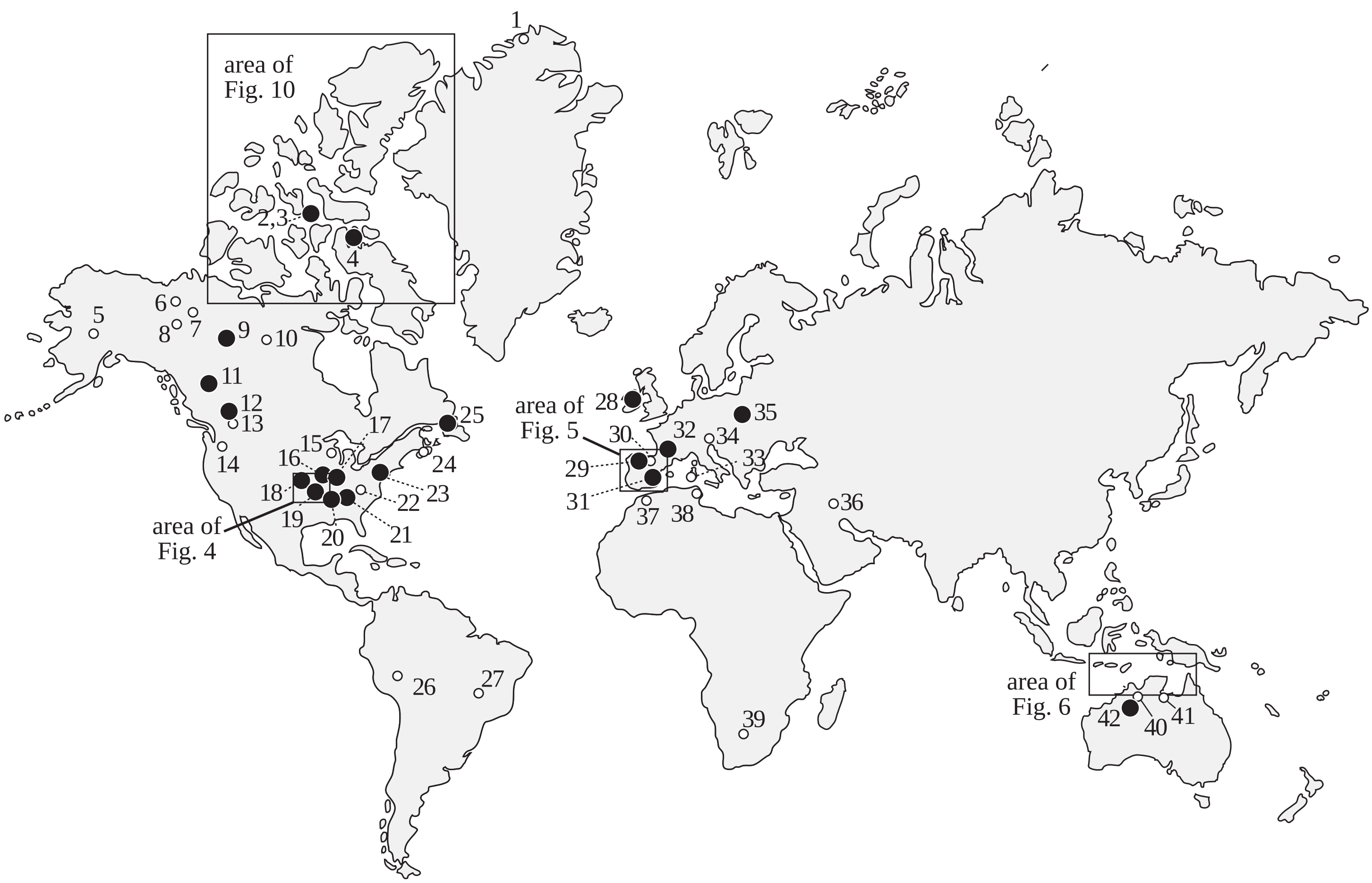


Figure 2, Bradley & Leach

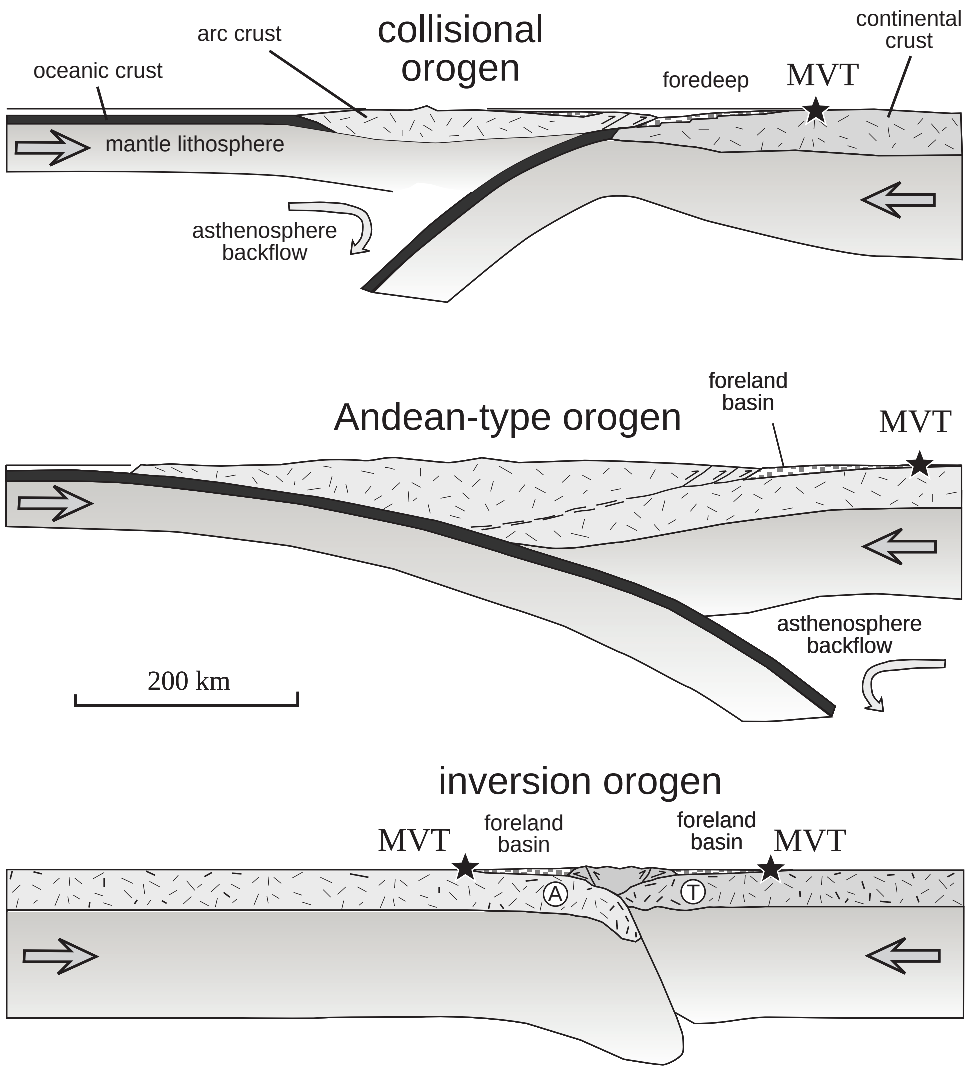


Figure 3, Bradley & Leach

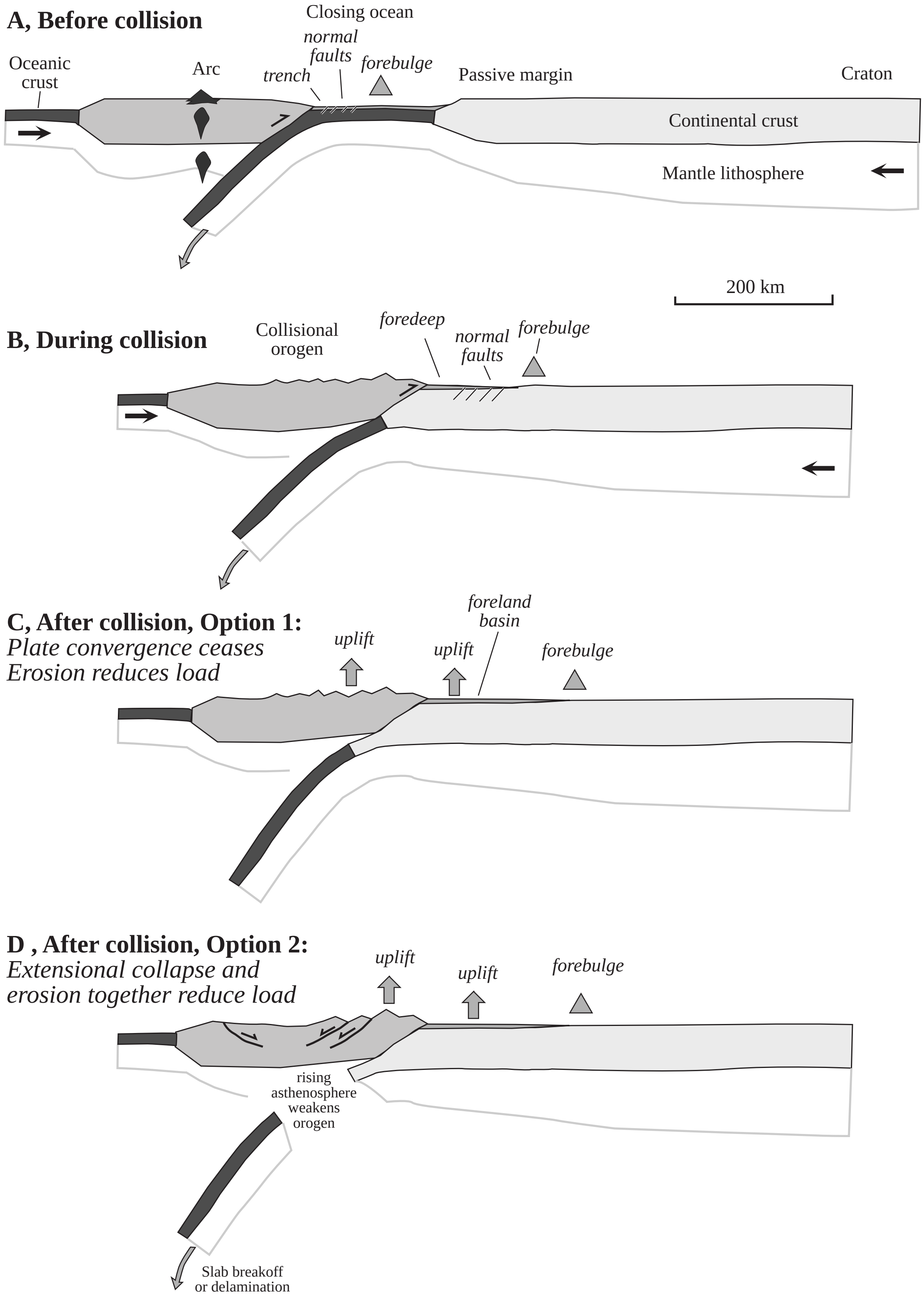
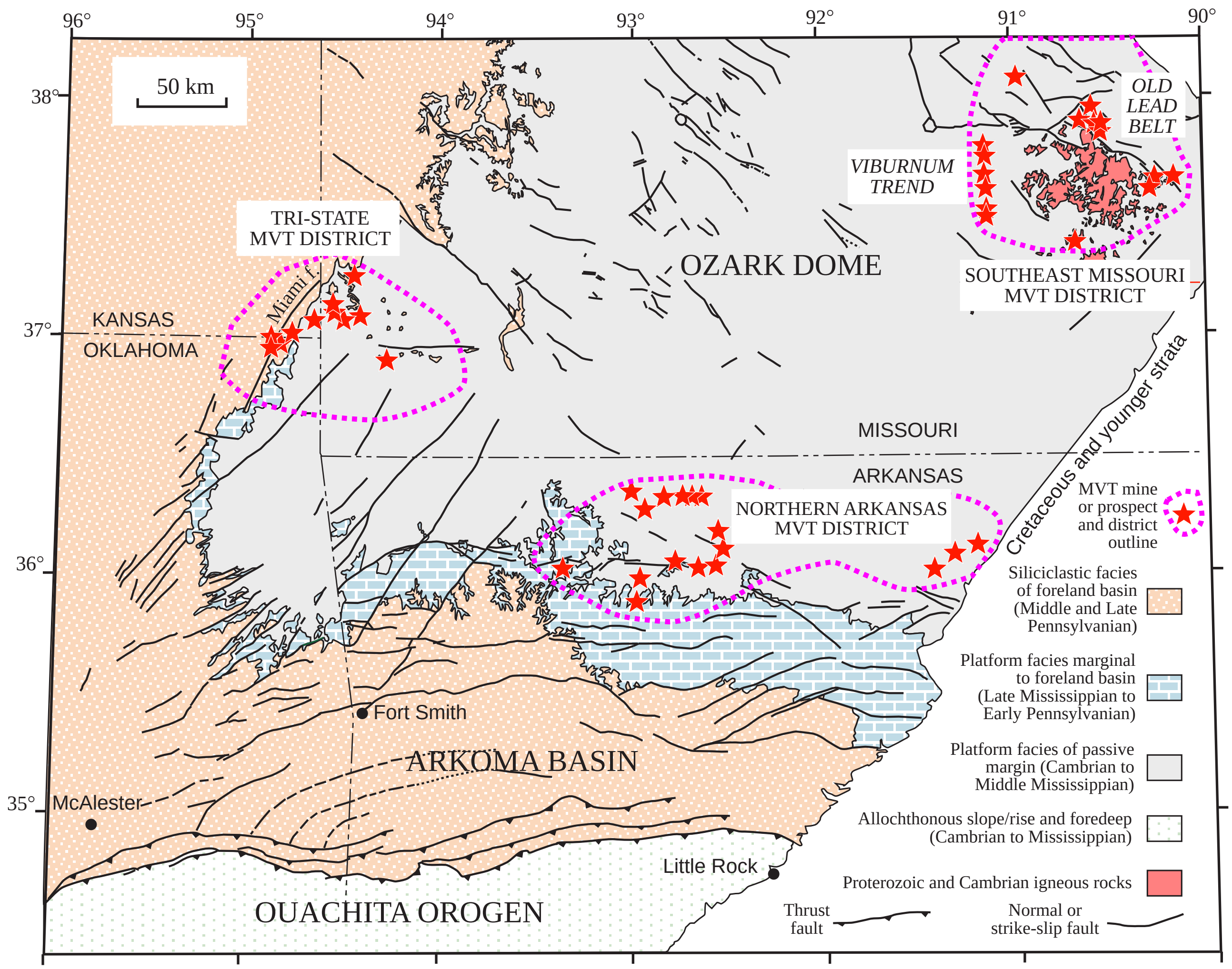


Figure 4, Bradley & Leach



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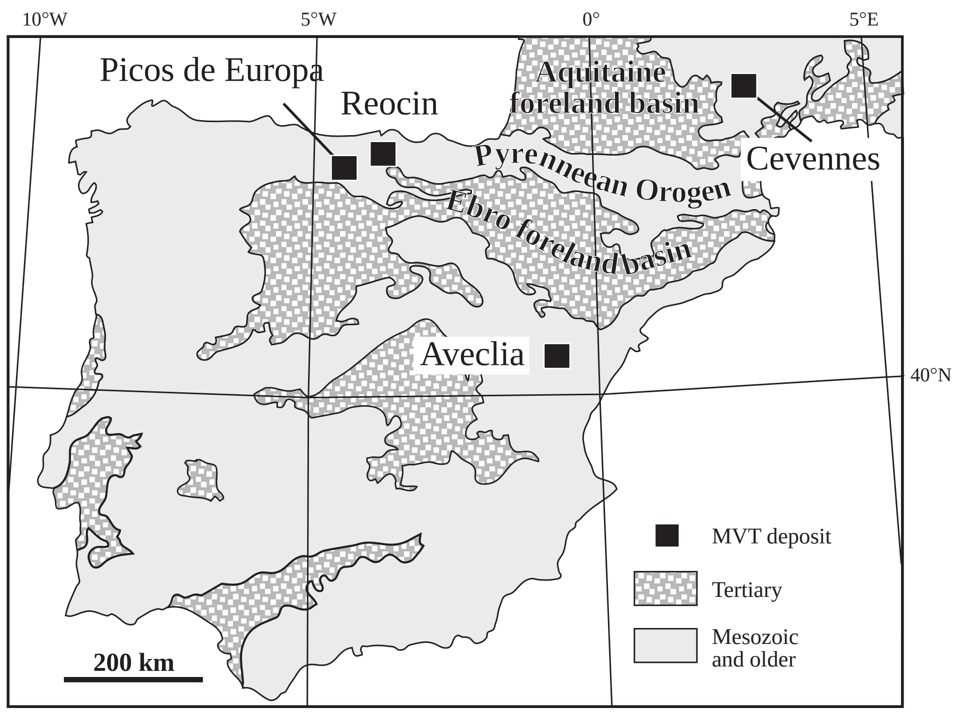

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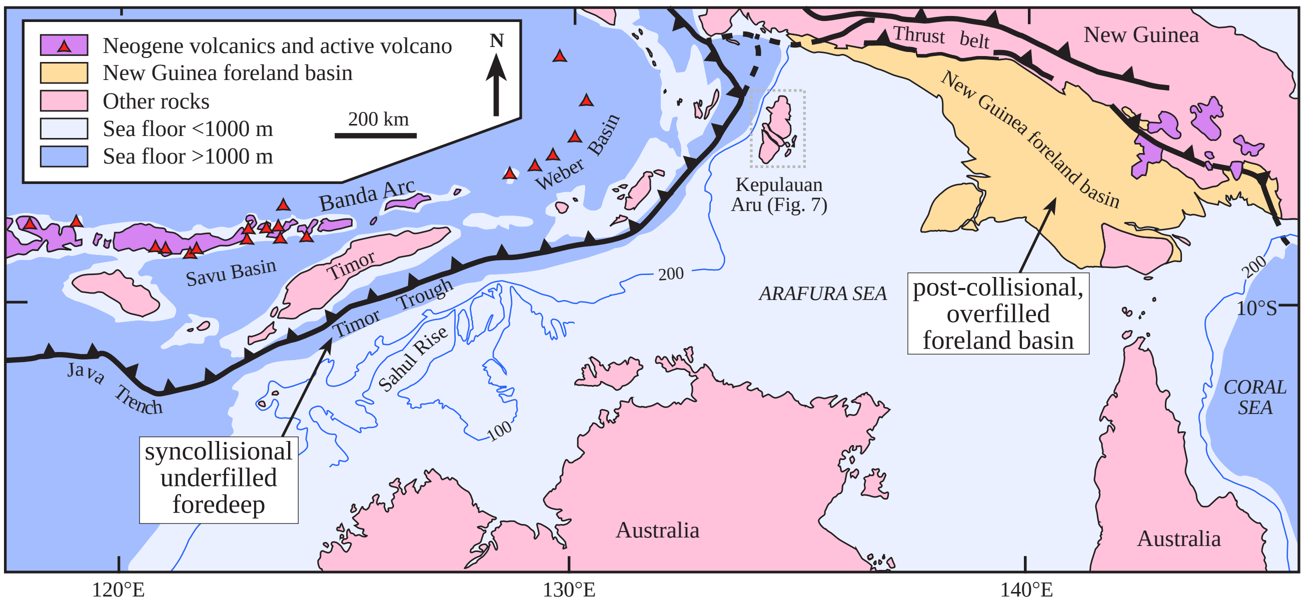

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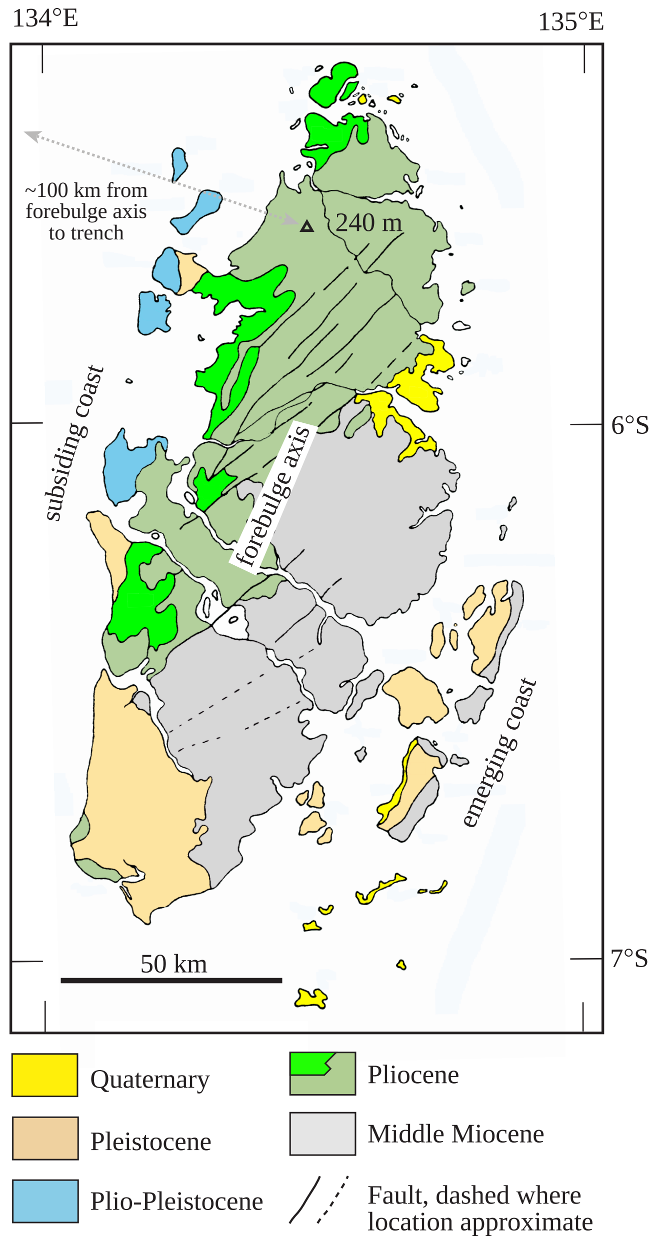

Figure 8, Bradley & Leach

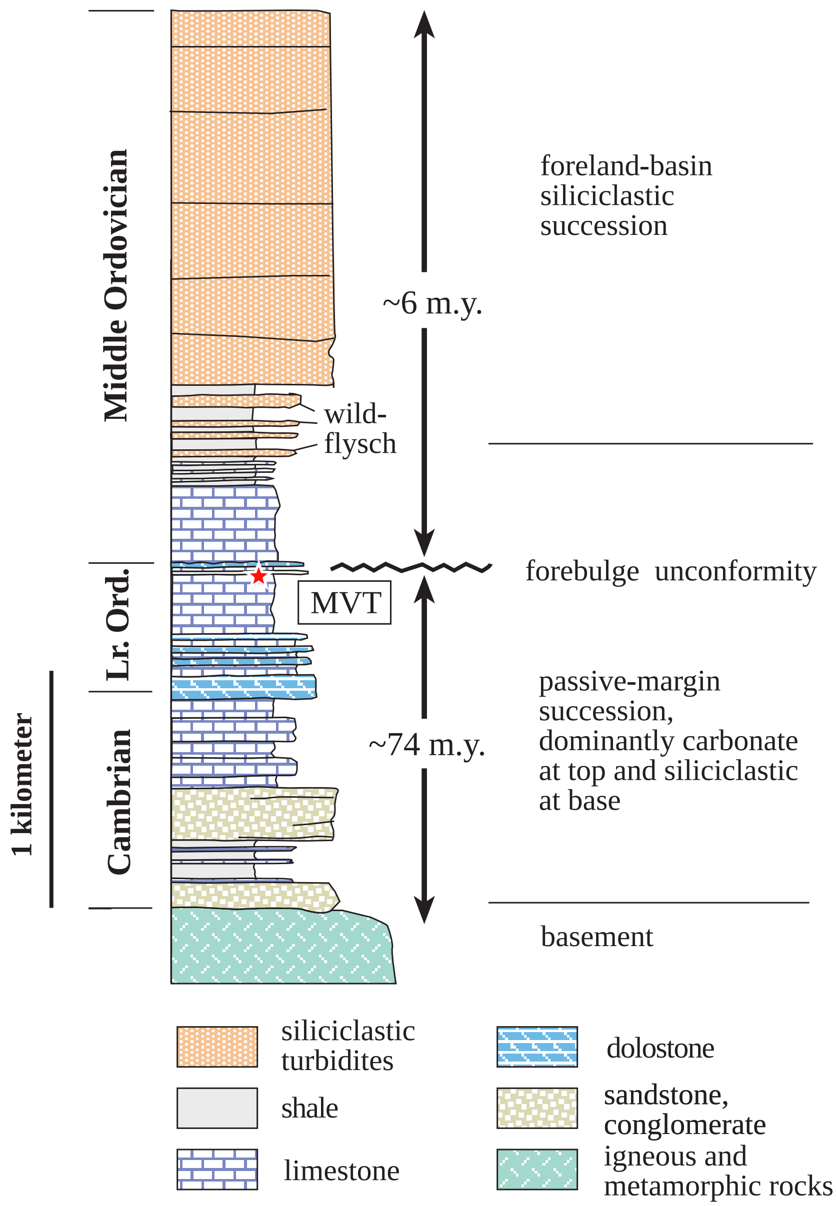

Figure 9, Bradley & Leach

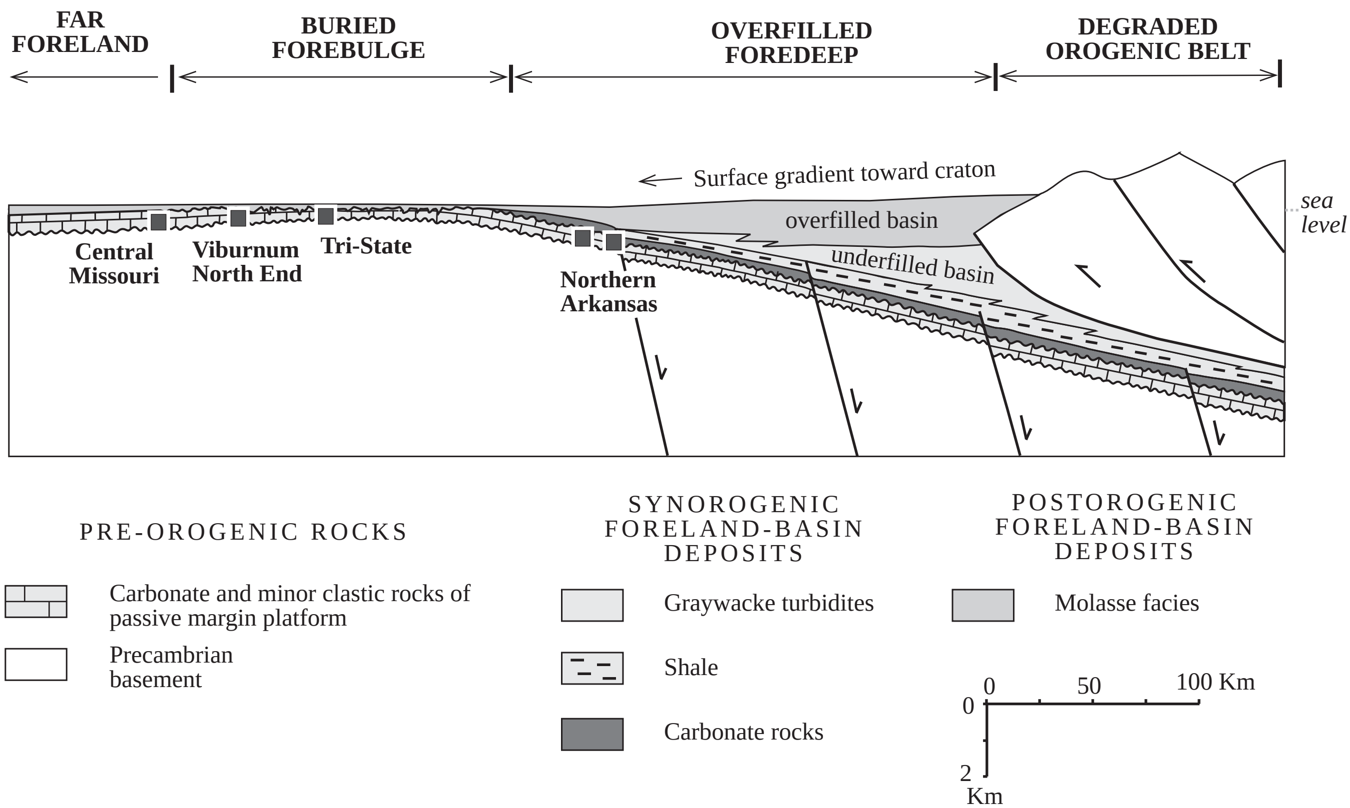

Figure 10, Bradley and Leach

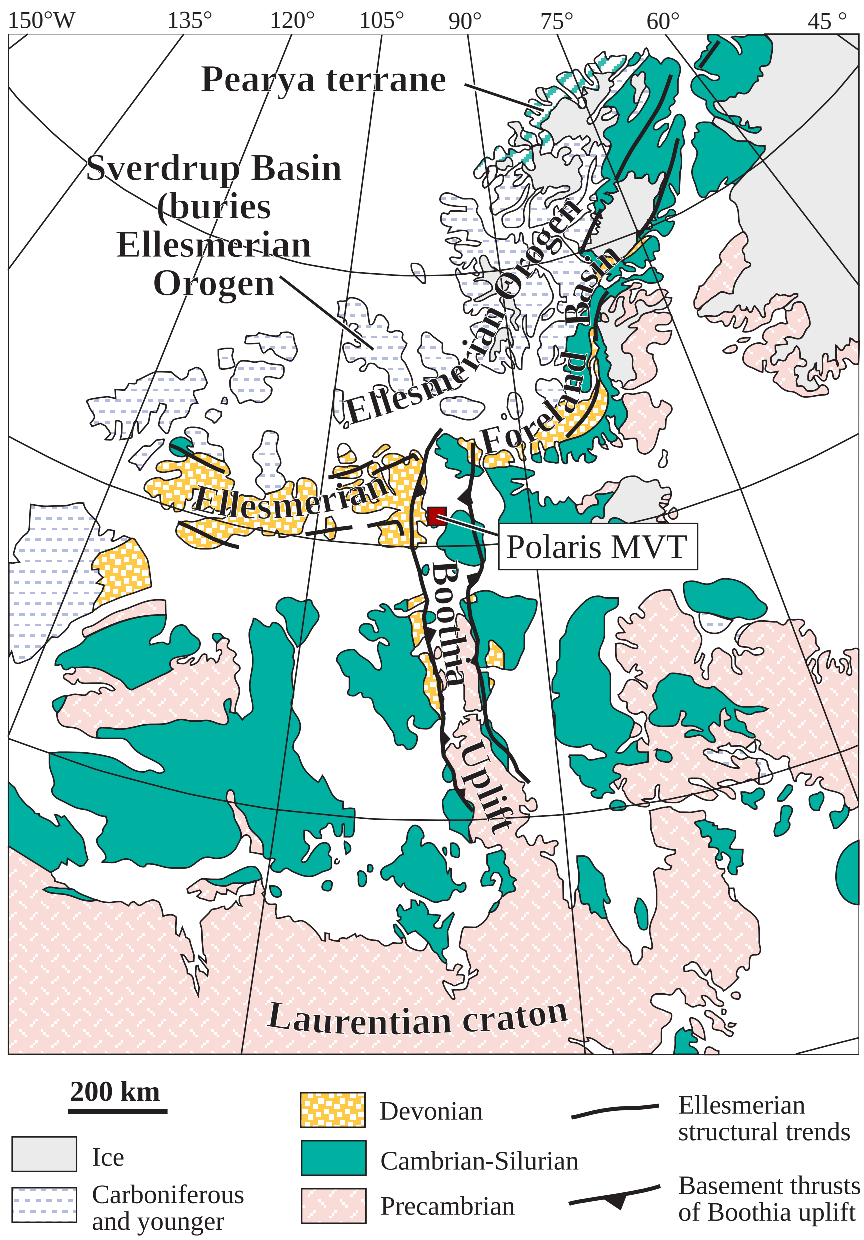

Figure 11, Bradley & Leach

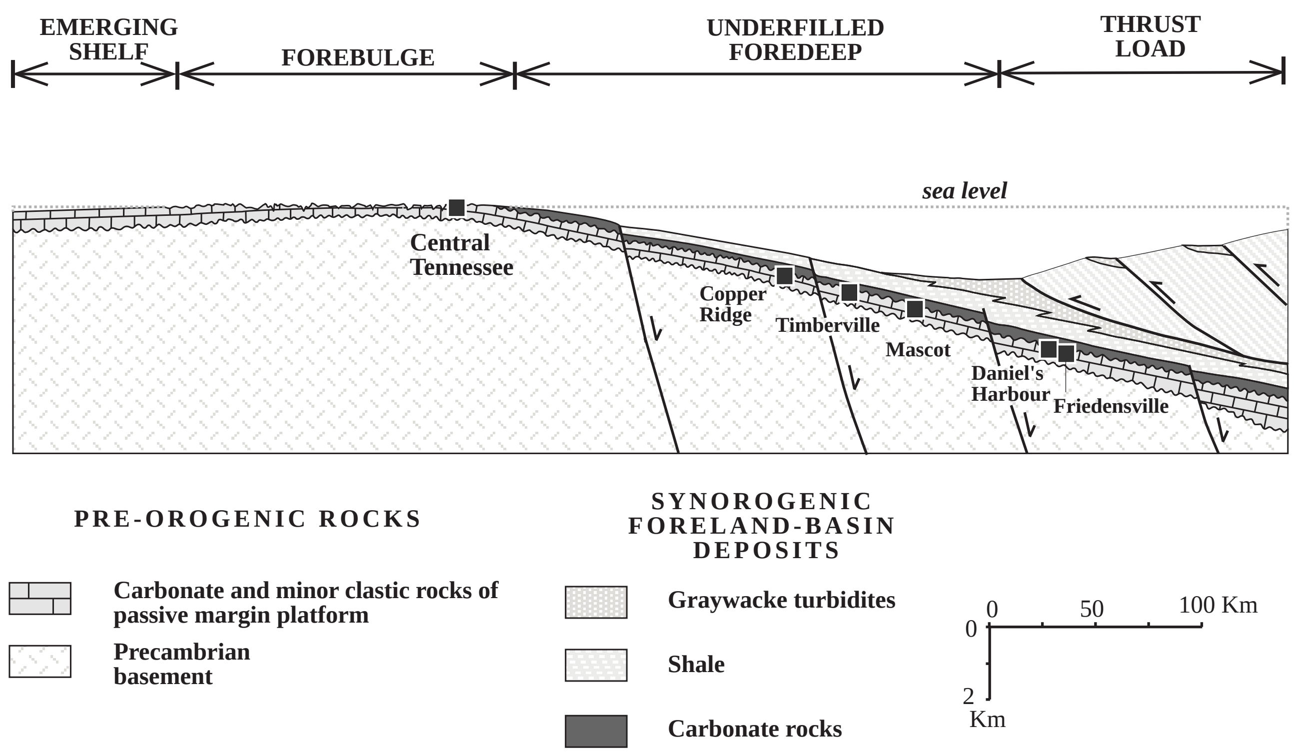

Figure 12, Bradley & Leach

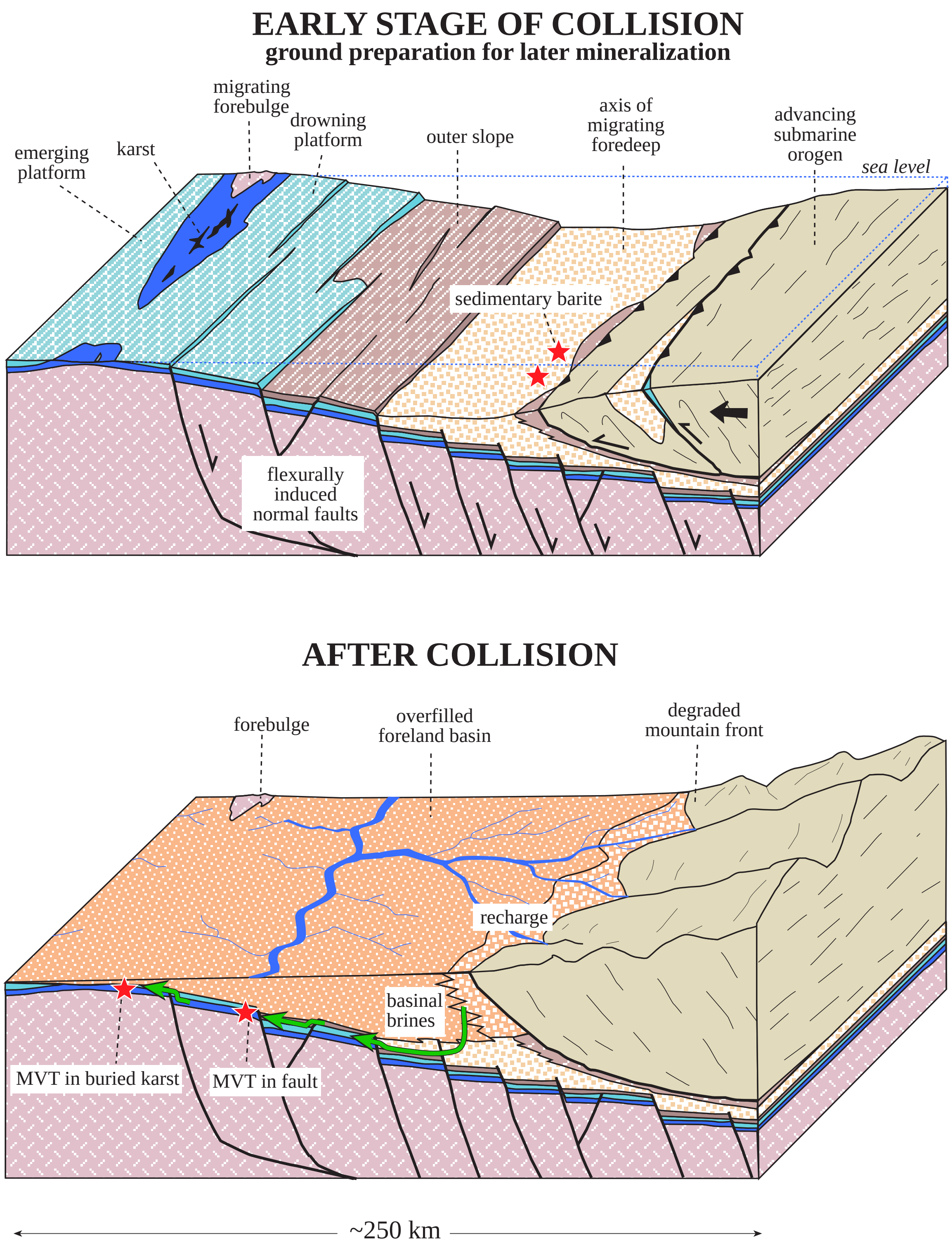

Figure 13, Bradley & Leach